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EVALUATION OF THE SIMPLE SETTLING THEORY FOR PREDICTING SEDIMENT DEPOSITION BY OVERLAND FLOW

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ABSTRACT

Experiments were conducted in the laboratory to evaluate the influence of sediment concentration, sediment grain-size distribution, bed slope and flow discharge on sediment deposition rates and patterns associated with a reduction in bed slope. The experimental data clearly indicate that sediment deposition by overland flow is a very selective process: fine particles remain almost entirely in suspension and coarse particles are deposited quickly. Analysis of the data shows that up to a critical unit discharge a simple settling equation without a transport term, assuming continuous mixing of the sediment and water, gives a good prediction of the overall sediment delivery ratio and the grain-size distribution of the deposited and the exported sediment. However, there are some discrepancies for the clay, the coarse silt and the sand fractions. The assumption of continuous mixing is tested by investigating the sedimentation patterns of very narrow size classes. The observed decrease of sediment concentration versus distance from the inflow point for these individual sediment size classes closely agrees with the prediction assuming continuous mixing. When the critical unit discharge is exceeded, hydraulic properties of the overland flow do influence the sediment delivery outcomes. At discharges exceeding the threshold value the simple settling theory underpredicts the sediment delivery ratio. In these hydraulic conditions, a transport term needs to be incorporated into the simple settling theory. It is shown that the transport capacity and the re-entrainment model yield similar expressions for the description of sediment transport by overland flow over an area of net deposition. The experimental data indicate that the re-entrainment of previously deposited sediment is non-selective. Copyright © 1999 John Wiley & Sons, Ltd.

KEY WORDS: sediment deposition; settling velocity; mixing; re-entrainment; overland flow

INTRODUCTION

Physically based soil erosion models necessarily contain a sediment deposition routine as they must be capable of routing sediment over areas of net deposition. Some of these deposition equations are based on the transporting capacity principle, e.g. CREAMS (Foster *et al.*, 1980), WEPP (Foster *et al.*, 1995), EUROSEM (Morgan *et al.*, 1998), LISEM (De Roo and Offermans, 1995). These models predict the local net deposition rate based on the difference between the transporting capacity of the flow and the actual sediment load on the one hand, and particle fall velocity on the other hand (Figure 1). The net deposition rate for sediment size class i, as for example incorporated in the WEPP model, can be written as (Foster, 1982):

$$D_{ri} = \beta v_{si}\left(\frac{Tc_i}{q}\frac{n_i}{n_{tot}} - \frac{c_i}{q}\right) \quad (1)$$

where D_{ri} = net deposition rate for size class i (kg s^{-1} m^{-2}), β = raindrop induced turbulence coefficient,

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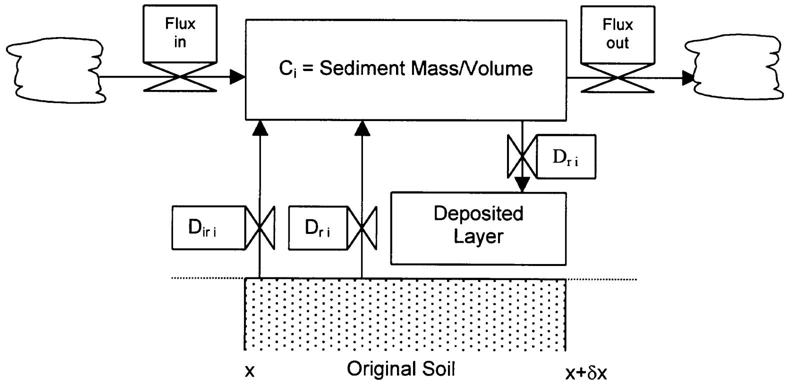

Figure 1. Flow diagram (after the style of Forrester 1970) describing the interaction of erosion processes between the sediment flux, the original soil and the deposited layer in the WEPP model. The process rates are interrill erosion (D_{iri}) and rill erosion/deposition (D_{ri})

v_{si} = effective fall velocity for sediment size class i (m s^{-1}), Tc_i = transport capacity of size class i in the rill (kg s^{-1} m^{-1}), c_i = sediment load of size class i (kg s^{-1} m^{-1}), q = unit discharge (m^2 s^{-1}), n_i = the number of transported particles of type i, n_{tot} = the total number of transported particles.

Hairsine and Rose (1992a,b) follow a different approach: in their model the local net deposition rate is calculated as the difference between gross deposition and re-entrainment (Figure 2). Net deposition is given by:

$$d_i - r_{ri} = \alpha_i C_i v_{si} - \frac{\alpha_i HF}{g}\frac{\sigma}{(\sigma-\rho)}\frac{(\Omega-\Omega_0)}{D}\frac{Mdi}{Mdt} \tag{2}$$

where d_i = mass rate of deposition per unit area of size class i (kg s^{-1} m^{-2}), r_{ri} = rate of sediment re-entrainment for size class i (kg s^{-1} m^{-2}), C_i = mean sediment concentration (size class i) (kg m^{-3}), α_i = ratio of the sediment concentration adjacent to the bed to the mean sediment concentration, ν_{si} = fall velocity of sediment size class i (m s^{-1}), H = fractional shielding of the soil by the deposited layer, F = fraction of stream power used for re-entrainment, g = gravity (m s^{-2}), σ = sediment density (kg m^{-3}), ρ = water density (kg m^{-3}), Ω = stream power (kg s^{-3}), Ω_o = critical stream power (kg s^{-3}), D = depth of the water flow (m), Mdi = mass of sediment of class i in the deposited layer (kg m^{-2}), Mdt = total mass of the deposited layer per unit area (kg m^{-2}).

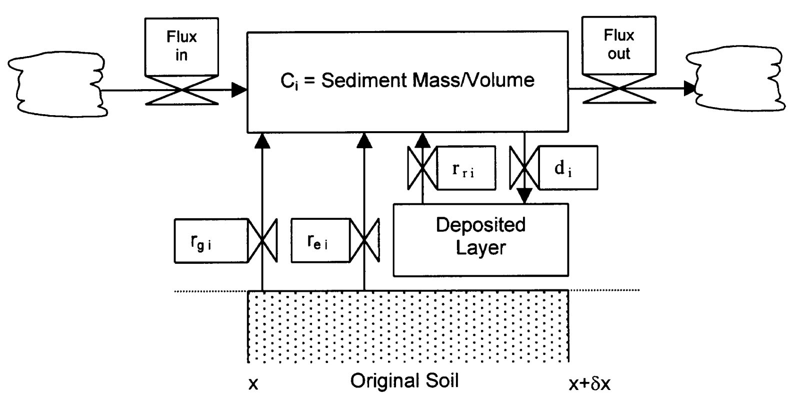

Figure 2. Flow diagram (after the style of Forrester 1970) describing the interaction of erosion processes between the sediment flux, the original soil and the deposited layer in the Hairsine–Rose model. The process rates are entrainment (r_{ei}), re-entrainment (r_{ri}), gravity processes (r_{gi}) and deposition (d_i) (Hairsine and Rose, 1992a)

Some models (LISEM, EUROSEM and the Hairsine–Rose model) incorporate a critical hydraulic threshold value in their formulation. At flow discharges lower than this threshold value, sediment transport or re-entrainment is assumed to be negligible. For this condition, both types of models can be reduced to a simplified model, predicting the deposition rate as a function of the sediment's fall velocity, the sediment concentration and the unit discharge. This model is referred to as the simple settling theory (SST). This model relies on several untested assumptions: (i) the fall velocity of a grain in a moving fluid is comparable to the fall velocity of the same grain in standing water, and (ii) there is a continuous, complete remixing of sediment and water, i.e. there is no vertical sediment concentration gradient within the flow. These assumptions are difficult to verify directly, considering the fact that overland flows are often only a few millimetres deep and heavily laden with sediment. Testing the SST has therefore to be carried out indirectly, by comparing predicted and observed grain-size distributions and sedimentation patterns.

A distinction can also be made between single-class and multi-class models. Single-class models predict sediment transport and deposition based on the mean grain size only (e.g. EUROSEM, LISEM). Single-class deposition theories can be useful to predict total sediment masses leaving a catchment and to follow the evolution of long-term landscape formation. The disadvantage of these models is that they do not give information about the size distribution of the deposited and exported sediment. The sediment size distribution is strongly related to the sediment quality, since pollutants are mainly sorbed to the finest particles. In multi-class models (e.g. CREAMS, WEPP and the Hairsine–Rose model), the sediment grain-size distribution is subdivided into a number of classes so that size selectivity in erosion/deposition processes can be taken into account. Considering the present-day focus on off-site effects of soil erosion (e.g. Slattery and Burt, 1997; Sharpley *et al.*, 1994) this is an important advantage.

Although a vast number of studies of different aspects of soil erosion processes at various spatial and temporal scales exist, there are very few detailed studies on sediment transport by overland flow over net deposition zones. Consequently, most of the equations describing depositional phases in physically based erosion–deposition models remain largely untested, although predictions of sediment delivery and sediment enrichment are very sensitive to the performance of the deposition equations.

A recent experimental study by Beuselinck *et al.* (in press) clearly indicates the necessity for a hydraulic threshold value in any deposition model. In this paper the results of these series of detailed flume experiments are used to evaluate the models described above. More specifically, the study investigates (i) to what extent sediment deposition (both total mass and sediment quality) can be predicted using the simple settling theory for flows with a shear stress below the threshold value, and (ii) which modelling approach can best be used once the threshold value is exceeded.

EXPERIMENTAL SET-UP

A detailed account of the experimental set-up and procedures is given by Beuselinck *et al.* (in press). In short, sediment deposition was studied in a 2·6 m long and 0·117 m wide flume, into which a water/sediment mixture was introduced via a lead-in flume. The flume used had a fixed bed. This ensured that samples were taken solely from the deposited layer and not from the parent material. Several other erosion/deposition studies (e.g. Proffitt *et al.* 1991; Moss *et al.* 1980) were conducted in a flume with a loose soil bed. In these experiments there is uncertainty about the thickness of the deposited layer. In our experiments this experimental uncertainty is avoided since no soil bed is involved.

The water/sediment mixtures were prepared by mixing known quantities of a silty soil material with a known quantity of water. Two soil materials with slightly different grain-size distributions, sampled at two different locations in the loam belt of central Belgium, were used (Table I). Both the incoming sediment concentration and flow discharge were predetermined. Experiments were carried out on three different slopes (1, 2 and 4 per cent) and unit flow discharges ranged from 0·00025 to 0·00170 $m^3 s^{-1}$ per metre width while incoming sediment concentrations ranged from 38 to 187 kg m^{-3}. In total, *c*. 40 experiments were carried out. Surface flow velocities were measured using dye tracing and were converted to mean flow velocities by using correction factors (Li *et al.*, 1996). Table II gives a summary of the range of flow hydraulic parameters measured. During each experiment the inflow and outflow was sampled for sediment concentration and

Table I. Grain-size distribution and standard deviation of the grain-size distribution of the two types of inflow sediment used

Size class	<2 μm	2–4 μm	4–8 μm	8–16 μm	16–32 μm	32–63 μm	>63 μm
Type A (mass %)	19·9	4·1	3·7	6·9	23·8	33·3	8·3
Type A, std dev.	2·43	0·43	0·67	1·01	0·87	0·99	0·79
Type B (mass %)	18·3	5·3	5·2	8·6	25	30·7	6·9
Type B, std dev.	1·15	0·47	0·51	0·75	0·44	2·19	1·08

Table II. Summary of hydraulic flow parameters

Hydraulic parameter	Range
Unit discharge	0·00024–0·00170 $m^2\ s^{-1}$
Mean flow velocity	0·26–0·53 m s^{-1}
Depth	0·6–4·3 mm

grain-size analysis. After the experiments, a detailed sampling of the sediment on the flume bed was carried out to assess the spatial distribution as well as the variation in grain-size characteristics of the deposited sediment. Data on the spatial distribution were used to calculate the decrease of sediment concentration in the flow with distance on the flume. All grain-size analyses were conducted with a Coulter LS-100 laser diffractometer. The procedures proposed by Beuselinck *et al.* (1998) were then used to convert these data to sieve-pipette data.

RESULTS

Beuselinck *et al.* (in press) showed that both the sediment delivery ratio (SDR), defined as the ratio of sediment delivered at the outlet to sediment introduced at the inlet of the flume, and the grain-size distribution of the exported sediment depend on hydraulic conditions as well as on the grain size distribution of the incoming sediment. They showed that a hydraulic threshold value, corresponding to a flow shear stress value of *c.* 0·61 Pa, exists below which the transport of grains > 32 μm is insignificant. Once the threshold value is exceeded, the export of grains > 32μm increases rapidly (Figure 3). On the other hand, fine particles (<16 μm) are easily transported and remain almost entirely in suspension. As can be seen in Figure 3, the delivery ratio below the threshold value varies most for the medium silt fraction (16–32 μm). At shear stresses below

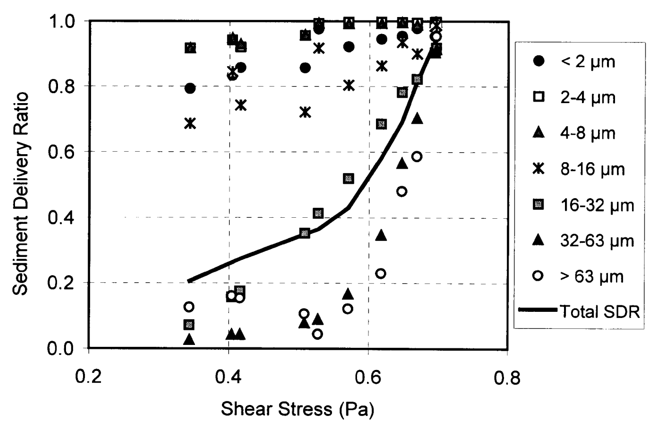

Figure 3. Sediment delivery ratio versus unit discharge for seven grain-size classes during experiments conducted on a 2 per cent slope with sediment type A and an inflow sediment concentration ranging from 75 to 85 kg m^{-3}

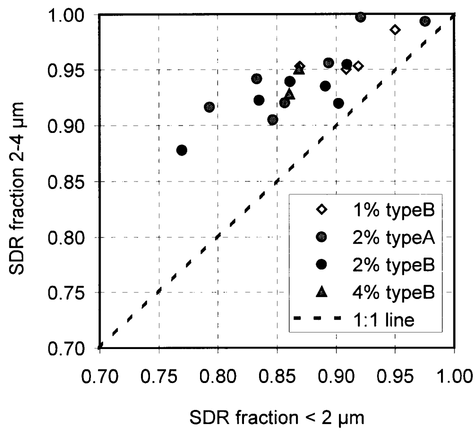

Figure 4. Sediment delivery ratio of the fine silt fraction (2–4 μm) versus the sediment delivery ratio of the clay fraction (< 2 μm) for experiments conducted on a 1 per cent, a 2 per cent and a 4 per cent slope with a unit discharge significantly lower than the threshold value (Beuselinck *et al.*, in press)

the threshold value, sediment delivery ratios for the fractions 2–4 μm and 4–8 μm are always somewhat higher than for the fraction < 2 μm (Figure 3 and 4). In similar hydraulic conditions, the SDR of the sand fraction depends on the slope at which deposition occurs (Figure 5). The SDR of the sand fraction is lower than the SDR of the coarse silt fraction (32–63 μm) on a 1 per cent slope, but is much higher on a 4 per cent slope.

DISCUSSION

The simple settling theory

The simplest model available to predict sediment deposition by overland flow in the absence of rainfall impact is the simple settling theory (SST) as described by Dabney *et al.* (1995). The multi-class SST assumes a steady flow that is uniform and a steady settling process in which particles settle without interference and are trapped when they reach the bed surface. Sediment re-entrainment of previously deposited sediment is

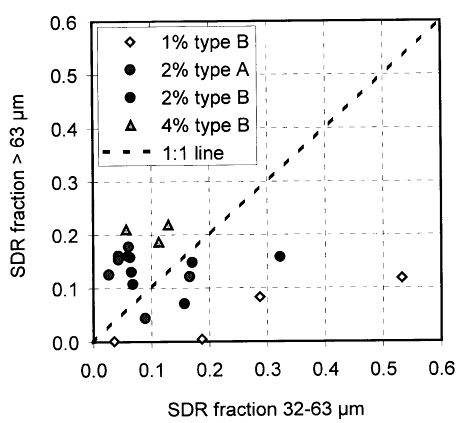

Figure 5. Sediment delivery ratio of the sand fraction (> 63 μm) versus the sediment delivery ratio of the coarse silt fraction (32–63 μm) for experiments conducted on a 1 per cent, a 2 per cent and a 4 per cent slope with a unit discharge significantly smaller than the threshold value (Beuselinck *et al.*, in press)

thus negligible. For this system the mass conservation equation is given by:

$$\delta C_i/\delta x = -(v_{si}/q)C_i \tag{3}$$

where C_i = local sediment concentration of fraction i (kg m^{-1}), x = distance (m), ν_{si} = settling velocity of fraction i (m s^{-1}), q = discharge per unit width (m^3 s^{-1} m^{-1}).

Equation 3 states that the deposition rate depends on the average remaining sediment load. Integrating, assuming continuous mixing of the sediment within the water, using the boundary condition that the local sediment concentration (C_i) equals the initial sediment concentration ($C_{i\ in}$) at $x = 0$, yields the fraction of each particle size class that reaches a distance, x, without settling:

$$C_i/C_{i\,in} = \exp(-(v_{si}/q)x) \tag{4}$$

If no mixing is assumed, the sediment concentration of a certain size class adjacent to the bed remains equal to the inflow sediment concentration of that size class until just before all particles of that size class are deposited. Integrating assuming that the local sediment concentration (C_i) equals the initial sediment concentration ($C_{i\ in}$) gives:

$$\begin{aligned} C_i/C_{i\,in} &= 1 - (v_{si}/q)x \text{ for } x \le q/v_{si} \\ C_i &= 0 \textit{ for } x > q/v_{si} \end{aligned} \tag{5}$$

It is noteworthy that Equations 4 and 5 do not contain a slope term. Furthermore, Equations 4 and 5 predict that the SDR is independent of the inflow sediment concentration.

The total SDR at the end of the flume is the sum of the fractions of each grain-size class reaching the outlet. The SDR versus distance in the case that mixing is assumed can thus be calculated by summing up a series of exponentially decreasing functions. On the other hand the SDR versus distance, when no mixing is assumed, can be obtained by summing up a series of linearly decreasing functions, each with a different slope gradient.

Transport term incorporated in erosion/deposition models

As two different approaches describing sediment transport over an area of net deposition exist, it is worthwhile to make a comparison.

At equilibrium (transport-limiting situation) re-entrainment for sediment class i in the Hairsine–Rose model equals deposition for sediment class i ($d_i - r_{ri}$, Equation 2) (Hairsine and Rose, 1992a):

$$\frac{\alpha_i HF}{g}\frac{\sigma}{(\sigma-\rho)}\frac{(\Omega-\Omega_0)}{D}\frac{Mdi}{Mdt} = \alpha_i v_{si} C \tag{6}$$

Hairsine and Rose (1992a) showed that at equilibrium:

$$\frac{Mdi}{Mdt} = \frac{v_{si}}{\sum_{i=1}^{I} v_{si}} \tag{7}$$

The limiting sediment concentration for size class i at equilibrium ($C_{eq\ i}$) then becomes (complete shielding, $H = 1$):

$$C_{eq\,i} = \frac{F}{g}\frac{\sigma}{(\sigma-\rho)}\frac{(\Omega-\Omega_0)}{D\sum_{i=1}^{I} v_{si}} \tag{8}$$

The expression for the rate of re-entrainment is (in case of complete shielding, $H = 1$):

$$r_{ri} = \frac{\alpha_i F}{g} \frac{\sigma}{(\sigma - \rho)} \frac{(\Omega - \Omega_0)}{D} \frac{Mdi}{Mdt} \tag{9}$$

Combining Equations 8 and 9 gives the general expression for the rate of re-entrainment:

$$r_{ri} = \alpha_i C_{eq\,i} \sum_{i=1}^{I} v_{si} \frac{Mdi}{Mdt} \tag{10}$$

The deposition rate of sediment size class i is described by:

$$d_i = \alpha_i v_{si} C_i \tag{11}$$

The net deposition rate (deposition minus re-entrainment: Equations 10 and 11) is therefore given by:

$$d_i - r_{ri} = -\alpha_i v_{si} \left(C_{eq\,i} \left(\frac{\frac{Mdi}{Mdt} \sum_{i=1}^{I} v_{si}}{v_{si}} \right) - C_i \right) \tag{12}$$

The deposition rate for sediment size class i as incorporated in the WEPP model can be written as (Foster, 1982):

$$D_{ri} = \beta v_{si} \left(\frac{Tc_i}{q} \frac{n_i}{n_{tot}} - \frac{c_i}{q} \right) \tag{1}$$

or

$$D_{ri} = \beta v_{si} \left(C_{eq\,i} \frac{n_i}{n_{tot}} - C_i \right)$$

with

$$C_{eq\,i} = \frac{Tc_i}{q}$$

and

$$C_i = \frac{c_i}{q} \tag{13}$$

Thus the multi-class deposition equations as incorporated in the WEPP and in the Hairsine–Rose model result in similar expressions for the description of sediment transport over an area of net deposition. The main differences are the conceptual approach to derive the sediment equilibrium concentration and the size selectivity of the deposition process. In the Hairsine–Rose model the *Mdi/Mdt* term is defined as the mass-ratio of particles of size class i in the total deposited layer (Hairsine and Rose, 1992a), whereas in the WEPP model the term n_i/n_{tot} is defined as the ratio of the numbers of particles of size class i in the flow and of the total number of particles in the flow (Foster, 1982). The transport term in the WEPP model (Tc_i) should be calculated separately for each particle class. An experimental verification of these concepts to describe the size selectivity of the deposition process has not yet been attempted.

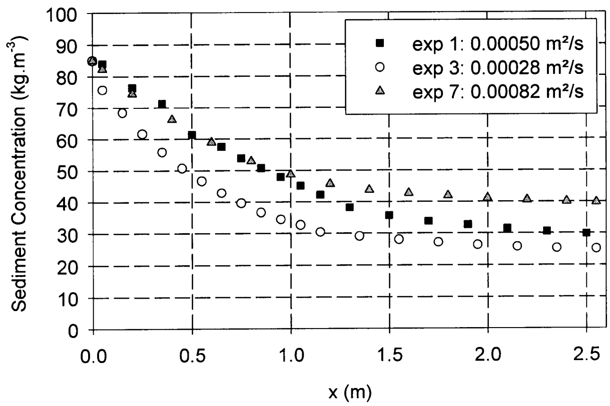

Figure 6. Sediment concentration versus downslope distance from the inflow point (x) for three experiments conducted on a 2 per cent slope with sediment type A and an inflow sediment concentration of *c*. 85 kg m^{-3}

Evaluation of the simple settling theory

An important assumption of the SST is that sediment mixing within the flow is perfect. This assumption has hitherto never been thoroughly tested for overland flow conditions. However, most deposition models assume that continuous mixing occurs (e.g. Storm *et al.*, 1994). This assumption is based on experimental data obtained by Meyer *et al.* (1995) and Einstein (1968). Meyer *et al.* (1995) studied the effect of vegetative barriers on sediment deposition. Einstein (1968) conducted experiments in a recirculating flume on deposition of suspended quartz particles (3–30 μm) in a river gravel bed. However, these experiments were carried out in deep, fully turbulent flows. It is questionable whether these results are valid for overland flow, which is often laminar to transitional. The Hairsine–Rose model (Hairsine and Rose, 1992a), on the other hand, incorporates a term, introduced by Croley (1982), to take into account the sediment concentration adjacent to the bed in the deposition routine (α_i). Croley (1982) states that sediment concentration has a non-uniform distribution with height above the bed. Sediment is assumed to mix uniformly throughout the water layer by rainfall impact in shallow water depths (Proffitt *et al.*, 1991).

In our experiments an exponential decrease of sediment concentration and thus deposition rate with distance downstream was also observed (Figure 6). However, this is not necessarily an indication that full mixing occurs. Figures 7 and 8 show the predicted variation of the total sediment concentration with downslope distance for one of our experiments in which the cases of no mixing and mixing are assumed (seven classes and 100 classes summed, respectively). Despite the linear decrease of sediment concentration for each individual size fraction, for no mixing occurring, the resulting curve closely matches a negative exponential decline: this is due to the fact that the linearly decreasing functions have different slope gradients and thus the individual sediment concentrations reach zero at different distances. If mixing is assumed, the resulting curve is also exponentially decreasing. If the grain-size distribution is separated into seven size classes (Figure 7) the total SDRs at the outlet are almost equal for both assumptions. If 100 size classes are summed (Figure 8) the SDR, assuming mixing, is significantly higher than if no mixing is assumed.

In order to evaluate whether mixing is present, it is necessary to investigate the sedimentation patterns of very narrow size classes. We used 0·155 size classes, corresponding to the smallest size classes measured by the Coulter LS-100. Figure 9 shows some typical results for two of the experiments (experiment 1: 0·00050 m^2 s^{-1}, 89 g l^{-1}; experiment 3: 0·00028 m^2 s^{-1} and 84 g l^{-1}). For the 24 and 48 μm particle sizes, the measured decrease of the sediment concentration of the individual sediment particle classes with distance agrees for both discharges much more closely with the exponential decline than with the linear decline, both predicted by using the settling velocity calculated with the Dietrich (1982) algorithm. For smaller particles (3 μm) the exponential decline is, owing to the higher sediment delivery of those particles, less obvious than for the coarser particles. These data thus clearly indicate that, independently of discharge, continuous mixing of the water and the sediment occurs in laminar to transitional flow conditions, even for coarse silt particles.

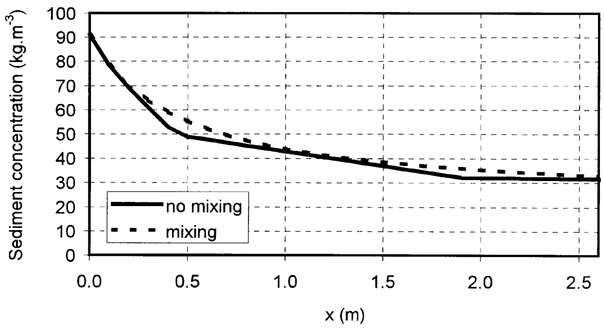

Figure 7. Total predicted sediment concentration (seven classes summed) versus distance (x) assuming mixing and no mixing

Therefore, in the following sections, full mixing is assumed. This shows that even without rainfall impact the term α_i in the Hairsine and Rose model (1992a), incorporated to account for a non-uniform vertical distribution in the flow, should be taken as unity.

Beuselinck *et al.* (in press) showed that a critical hydraulic shear stress exists above which significant sediment transport occurred concurrently with sediment deposition (Table III). The SST was therefore evaluated using sediment deposition data for which hydraulic conditions were clearly below this threshold value. The sediment delivery ratio was calculated for the seven size classes shown in Table I. Fall velocities for each size fraction were calculated using the equations developed by Dietrich (1982) using a shape factor of 0·7 and a roundness value of 3·5.

Despite sediment accumulation and related hydraulic changes, the outflow sediment concentration remains almost constant during the experiment (Figure 10). Since the increase of the slope of the sediment mass is rather limited, the threshold value (Beuselinck *et al.*, in press) for transport/re-entrainment of previously deposited sediment, which is slope-dependent, is not exceeded during an experiment. As long as the threshold value is not exceeded, the sediment outflow concentration remains constant.

Figure 11 shows that the SST predicts the observed overall SDR reasonably well. The observed independence of the SDR from the inflow sediment concentration is in agreement with the predictions of the SST. The SST also correctly predicts a higher SDR for sediment type B. This indicates that small changes in particle size distribution can have major effects on the predicted and measured SDR, as described by Dabney *et al.* (1995). This is due to the non-linear dependence of fall velocity on particle size (Graf, 1971). However, there are also clear discrepancies. In general, there is an overestimation of low SDRs and an underestimation of SDRs higher than 0·4. Furthermore, in contrast with theoretical predictions, observed SDRs appear to be somewhat dependent on the slope gradient (Beuselinck *et al.*, in press).

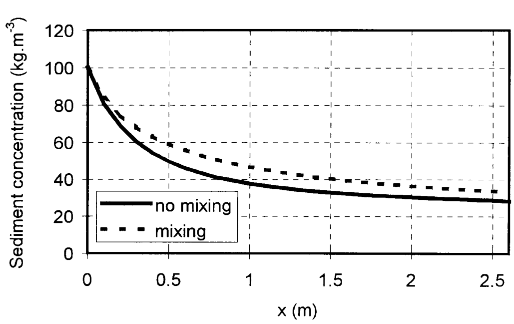

Figure 8. Total predicted sediment concentration (100 classes summed) versus distance (x) assuming mixing and no mixing

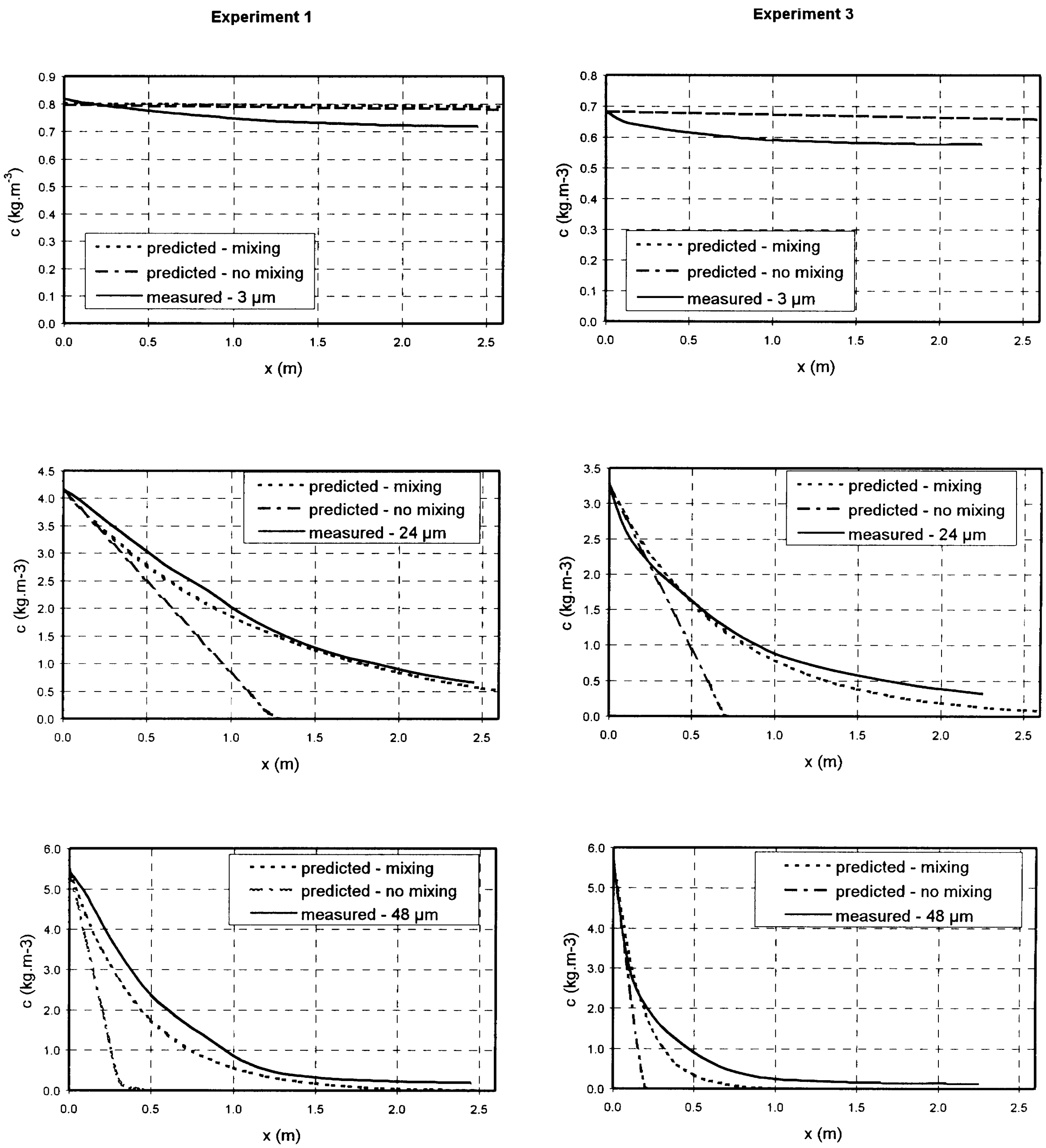

Figure 9. Predicted (assuming mixing and no mixing) and measured sediment concentration (c) for individual size classes of 0·155ϕ (3, 24 and 48 μm) versus distance (x) (for experiments 1 and 3)

If the SDR is calculated separately for each size class, it can be seen that there is a good overall agreement between the observed and predicted SDR for the various size classes, indicating that SST is capable of predicting both total sediment output and size selectivity in this domain (Figure 12). However, the SST underpredicts the fractions coarser than 16 μm. Underprediction is strongest for the sand and coarse silt fraction (32–63 μm). On the contrary, there is an overprediction of the clay (< 2 μm) export.

The small under- and overprediction of the SDR by the SST (Figure 12) can be attributed to several sources. A first explanation is the use of a limited number of size classes, corresponding to the phi-classes.

Table III. Critical unit flow discharges and critical flow shear stresses for data obtained at slopes ranging from 1 to 8 per cent (Beuselinck et al., in press)

Slope (m m^{-1})	Critical unit discharge (m^2 s^{-1})	Critical shear stress (Pa)
0·01	0·00205	0·56
0·02	0·00096	0·58
0·03	0·00059	0·58
0·04	0·00039	0·62
0·06	0·00026	0·71
0·08	0·00010	0·62
Average std dev.		0·61
		0·05

Predicting the sediment delivery ratio using smaller size classes will give a more accurate prediction of the SDR. Ideally an unlimited number of size classes should be used. Secondly, the settling velocity is calculated with an algorithm presented by Dietrich (1982), based on assumptions of quiescent settling conditions in turbulent free flow. However, Storm *et al.* (1994) cite studies by Chen (1975), Brown (1950), Camp (1946) and Dobbins (1944) which state that turbulent flow, which is often achieved under field conditions, influences settling velocities and reduces deposition rate. Thirdly, the settling velocity is also affected by the particle concentration in the flow. It is considered that the settling velocity decreases with increasing sediment concentration due to flow distortion by other settling particles (Allen, 1985).

The processes described above may be responsible for a general over- or underprediction of the SDR, but can hardly explain the strong underprediction of the coarse silt and the sand export and the overprediction of the clay export as they will affect all size fractions. The consistent overprediction of the clay export, which was also noted by Dabney *et al.* (1995), can possibly be explained by the flocculation or coagulation of fine sediment, the trapping of fine sediment out of the flow by coarser particles and/or the effects of Brownian motion, which is especially significant for smaller particles (Dabney *et al.*, 1995; Lick, 1982). In order to examine which process causes the trapping of fine sediment, an additional deposition experiment was carried out using an artificial milled quartz sediment that is cohesionless and does not flocculate ($D_{50} = 32$ μm). The experimental data show that the SDR of the fraction < 2 μm is significantly lower than the SDR of the 2–4 μm fraction (Table IV), as observed in the experiments carried out with silty soil material. This indicates that trapping of fines by coarser sediment is important. This is in agreement with observations of increased settling velocities in a polydispersed suspension by Lovell and Rose (1991). The underprediction of the export of coarse sediment (> 32 μm) may be explained by bedload transport and by so-called uncommon selectivity (Poesen and Savat, 1980; Savat, 1982; Govers, 1989): as the flow in this domain was laminar to transitional, deposited sand particles may be more easily transported by rolling over the bottom of the flume than the finer fractions. Rolling of sand grains was observed during the experiments. Bedload transport is

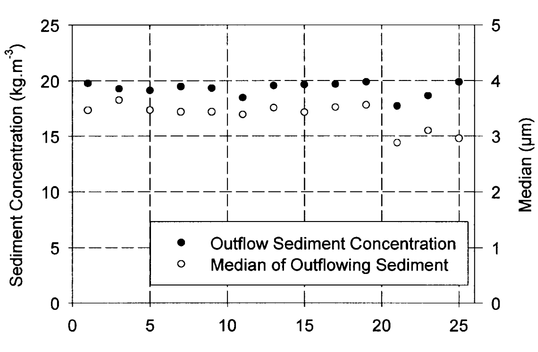

Figure 10. Outflow sediment concentration and median of the exported sediment versus time for an experiment conducted on a 2 per cent slope with sediment type A, an inflow sediment concentration of *c*. 80 kg m^{-3} and a unit discharge of 0·0003 m^2 s^{-1}

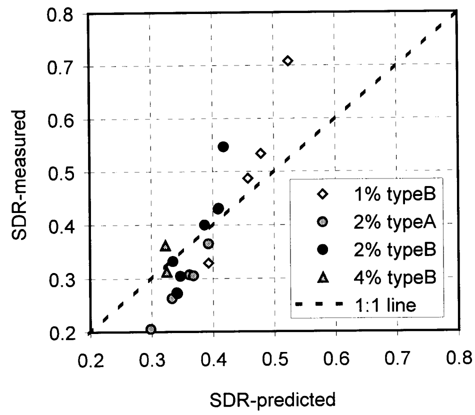

Figure 11. Predicted (simple settling theory) versus measured overall sediment delivery ratio for experiments conducted with a unit discharge clearly lower than the threshold value (Beuselinck *et al.*, in press) and on a 1 per cent, a 2 per cent and a 4 per cent slope

assumed to increase with slope gradient (e.g. Graf, 1971), which corresponds well with the observation that the SDR of the sand fraction, and to a lesser extent the overall SDR, increases with slope (Figure 5).

Bedload transport below the threshold value also contributes to the underprediction of sediment delivery ratios higher than 0·4 by the SST. The SDR calculated by using the measured SDR for the clay and for the coarse silt fractions and the SST theory for the other fractions agrees much more closely with the measured SDR (Figure 13). This clearly shows that the overestimation by the SST of SDR lower than 0·4 and the underestimation of SDR higher than 0·4 is respectively caused by the strong overprediction of the clay export and the presence of bedload transport of the coarse silt fraction at discharges higher than 0·0004 $m^2 s^{-1}$.

Evaluation of the transport term incorporated in erosion/deposition models

Figure 3 shows that the sediment delivery ratio of the 32–63 μm and > 63 μm fractions increases very sharply once the threshold value is exceeded, resulting in export of much coarser sediment than below the threshold value. In such conditions more sediment is exported than predicted by the SST. The SST would predict export of less coarse sediment than observed once the threshold discharge is exceeded. This implies that a significant amount of coarse sediment is transported by the flow over the depositional area. In such conditions a transport term needs to be incorporated in the deposition routine.

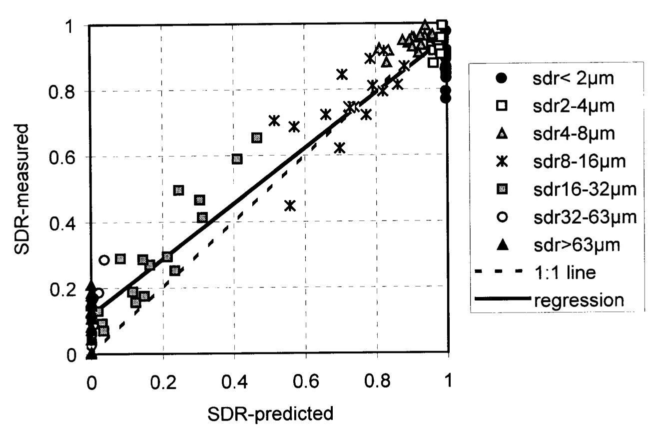

Figure 12. Predicted (simple settling theory) versus measured sediment delivery ratio for seven grain size classes for experiments conducted with unit discharges significantly lower than the threshold value (Beuselinck *et al.*, in press) and on a 1 per cent, a 2 per cent and a 4 per cent slope with a single regression line fitted through all the data

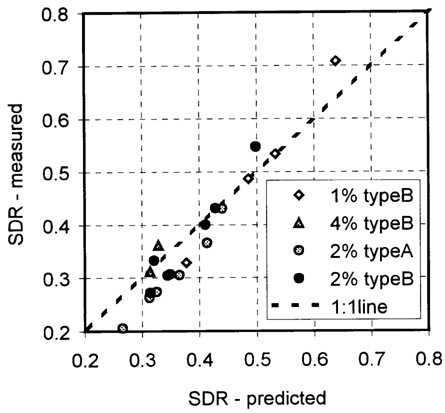

Figure 13. Predicted-measured (measured SDR for clay and coarse silt fraction (32–63 μm) and simple settling theory for the other fractions) versus measured overall sediment delivery ratio for experiments conducted with a unit discharge significantly lower than the threshold value (Beuselinck *et al.*, in press) and on a 1 per cent, a 2 per cent and a 4 per cent slope

The Hairsine–Rose model (Equation 2) predicts that sediment re-entrainment is unselective, i.e. the re-entrainment rate for each size class is proportional to the mass fraction of the size class in the deposited sediment. Figure 14 shows the observed grain-size distribution of that part of the exported sediment that cannot be explained by the settling of the particles, which can be interpreted as the re-entrained sediment. For this experiment, conducted at a discharge well above the threshold value, it is shown that the grain-size distribution of the re-entrained sediment is similar to the grain-size distribution of the deposited sediment. This indicates that in this discharge range unselective re-entrainment of previously deposited sediment occurs. The observed sediment sorting during sediment transport over an area of net deposition is thus only due to selective settling of the particles.

CONCLUSIONS

Sediment transport by overland flow across an area of net deposition can be seen as a combination of transport of suspended load, which settles according to the particle fall velocity, assuming a constant mixing of sediment and water, and re-entrainment of previously deposited sediment. Sediment re-entrainment is limited at lower discharges. The simple settling theory, predicting the settling of the suspended load, can be used to model deposition at these low flow rates if information concerning topography, hydraulic conditions and sediment composition are known. The observed independence of the SDR from the inflow sediment concentration and the influence of the grain-size distribution of the inflow sediment on the deposition process are in agreement with the predictions of the simple settling theory. However, there are some discrepancies, especially for the coarse silt, the sand and the clay fractions, resulting in overprediction of the clay export and underprediction of the coarse silt and the sand export.

It is important to stress that application of the simple settling theory requires knowledge of the vertical distribution of suspended sediment within the flow. Our experiments show that, even in the absence of rainfall, there is a continuous mixing of water and sediment, resulting in a vertically uniform sediment distribution for all size fractions < 50 μm.

Table IV. Sediment delivery ratios (SDR) for each size class; experiment carried out with a milled quartz sediment, on a 2 per cent slope, with a discharge of 0·00069 m^2 s^{-1} and an inflow sediment concentration of c. 80 kg m^{-3}

Size class	<2 μm	2–4 μm	4–8 μm	8–16 μm	16–32 μm	32–63 μm	>63 μm	Total
SDR	0·64	0·80	0·87	0·76	0·41	0·11	0·03	0·29

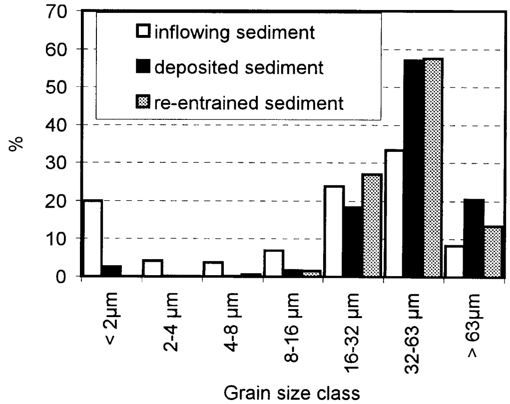

Figure 14. Comparison of the grain-size distribution of the 'transported' sediment (difference between the grain size of the measured exported sediment and the grain size of the predicted exported sediment using the simple settling theory) and the deposited sediment for an experiment conducted on a 2 per cent slope with sediment type A, an inflow sediment concentration of 80 kg m^{-3} and a unit discharge of 0·0012 $m^2 s^{-1}$

There is a sharp rise in re-entrainment of the deposited material once a threshold value is exceeded. In such conditions, a transport term has to be incorporated in the deposition equation. Mathematically, the deposition routines, based on transport capacity and on re-entrainment of previously deposited sediment, incorporated in erosion/deposition models result in similar expressions. The main difference between the two model types is the way the size selectivity is expressed. The term *Mdi/Mdt* incorporated in the Hairsine–Rose model seems to be a valuable term in predicting sediment sorting in zones of net deposition.

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REFERENCES

Allen, J. R. L. 1985. *Principles of Physical Sedimentology*, Allen & Unwin, London.

Beuselinck, L., Govers, G., Poesen, J., Degraer, G. and Froyen, L. 1998. 'Grain size analysis by Coulter LS-100: comparison with the sieve-pipette method', *Catena*, **32**, 193–208.

Beuselinck, L., Govers, G., Steegen, A. and Quine, T. A. (in press). 'Sediment transport by overland flow over an area of net deposition', *Hydrol. Process*.

Brown, C. B. 1950. 'Sediment transportation', in Rouse, H. (Ed.), *Engineering Hydraulics*, John Wiley, New York.

Camp, T. R. 1946. 'Sedimentation and design of settling tanks', *Trans. ASCE*, **111**, 895–958.

Chen, C. 1975. 'Design of sediment retention basins', in *Proceedings of National Symposium on Urban Hydrology and Sediment Control*, UK BU 109, College of Engineering, University of Kentucky, Lexington.

Croley, T. E. 1982. 'Unsteady overland flow sedimentation', *J. Hydrol.*, **56**, 325–346.

Dabney, S. M., Meyer, L. D., Harmon, W. C., Alonso, C. V. and Foster, G. R. 1995. 'Depositional patterns of sediment trapped by grass hedges', *Trans. ASAE*, **38**(6), 1719–1729.

De Roo, A. P. J. and Offermans, R. J. E. 1995. 'LISEM: a physically-based hydrological and soil erosion model for basin scale water and sediment management', in *Modeling and Management of Sustainable Basin-scale Water Resource Systems (Proceedings of a Boulder Symposium, July 1995)*, IAHS Publication No. 231, 399–407.

Dietrich, W. E. 1982. 'Settling velocity of natural particles', *Water Resour. Res.*, **18**(6), 1615–1626.

Dobbins, W. E. 1944. 'Effect of turbulence on sedimentation', *Trans. ASCE*, **109**, 619–678.

Einstein, H. A. 1968. 'Deposition of suspended particles in a gravel bed', *J. Hydraulic Div., Proc. ASCE*, **94**(HY5), 1197–1205.

Forrester, J. E. 1970. *Industrial Dynamics*, MIT Press, Cambridge, Mass.

Foster, G. R. 1982. 'Modeling the erosion process', in Haan, C. T. (Ed.), *Hydrologic Modeling of Small Watersheds*, American Society of Agricultural Engineers, ASAE Monograph, No. 5, St. Joseph, Mich., 297–360.

Foster, G. R., Lane, L. J., Nowlin, J. D., Laflen, J. M. and Young, R. A. 1980. 'A model to estimate sediment yield from field-sized areas: development of model', in Knisel, W. G. (Ed.), *CREAMS – A Field-scale Model for Chemicals, Runoff and Erosion from Agricultural Management Systems*, USDA Cons. Res. Report No. 26, USDA-SEA, 36–64.

Foster, G. R., Flanagan, M. A., Nearing, M. A., Lane, L. J., Risse, L. M. and Finkner, S. C. 1995. 'Hillslope erosion component', in Flanagan, D. C. and Nearing, M. A. (Eds), *USDA Water Erosion Prediction Project: Hillslope profile and Watershed Model Documentation*, NSERL Report No. 10, USDA-ARS National Soil Erosion Research Laboratory, West Lafayette, Ind., 11 pp.

Govers, G. 1989. 'Grain velocities in overland flow: a laboratory study', *Earth Surf. Process. Landforms*, **14**, 481–498.

Graf, W. H. 1971. *Hydraulics of Sediment Transport*, McGraw-Hill, New York.

Hairsine, P. B. and Rose, C. W. 1992a. 'Modeling water erosion due to overland flow using physical principles 1.) Sheet flow', *Water Resour. Res.*, **28**(1), 237–243.

Hairsine, P. B. and Rose, C. W. 1992b. 'Modeling water erosion due to overland flow using physical principles 2.) Rill flow', *Water Resour. Res.*, **28**(1), 245–250.

Li, G., Abrahams, A. D. and Atkinson, J. F. 1996. 'Correction factors in the determination of mean velocity of overland flow', *Earth Surf. Process. Landforms*, **21**, 509–515.

Lick, W. 1982. 'Entrainment, deposition and transport of fine-grained sediment in lakes', *Hydrobiologia*, **91**, 31–40.

Lovell, C. J. and Rose, C. W. 1991. 'Wake-capture effects observed in a comparison of methods to measure particle settling velocity beyond Stokes'range', *J. Sed. Petrol.*, **61**(4), 575–582.

Meyer, L. D., Dabney, S. M. and Harmon, W. C. 1995. 'Sediment-trapping effectiveness of stiff-grass hedges', *Trans. ASAE*, **38**(3), 809–815.

Morgan, R. P. C., Quinton, J. N., Smith, R. E., Govers, G., Poesen, J. W. A., Auerswald, K., Chisci, G., Torri, D. and Styczen, M. E. 1998. 'The European Soil Erosion Model (EUROSEM): a dynamic approach for predicting sediment transport from fields and small catchments', *Earth Surf. Process. Landforms*, **23**, 527–544.

Moss, A. J., Walker, P. H. and Hutka, J. 1980. 'Movement of loose, sandy detritus by shallow water flows: an experimental study', *Sed. Geol.*, **25**, 43–66.

Poesen, J. and Savat, J. 1980. 'Particle-size separation during erosion by splash and run-off', in De Boodt, M. and Gabriels, D. (Eds), *Assessment of Erosion*, Wiley, Chichester, 427–439.

Proffitt, A. P. B., Rose, C. W. and Hairsine, P. B. 1991. 'Rainfall detachment and deposition, I. Experiments with low slopes and significant water depths', *Soil Sci. Soc. Am. J.*, **55**(2), 325–332.

Savat, J. 1982. 'Common and uncommon selectivity in the process of fluid transportation: Field observations and laboratory experiments on bare surfaces', in Yaalon, D. H. (Ed.), *Catena* Supplement 1, 139–160.

Sharpley, A. N., Chapra, S. C., Wedepohl, R., Sims, J. T., Daniel, T. C. and Reddy, K. R. 1994. 'Managing agricultural phosphorus for protection of surface waters: issues and options', *J. Environ. Qual.*, **23**, 437–451.

Slattery, M. C. and Burt, T. P. 1997. 'Particle size characteristics of suspended sediment in hillslope runoff and stream flow', *Earth Surf. Process. Landforms*, **22**, 705–719.

Storm, D. E., Barfield, B. J. and Altenhof, C. T. 1994. 'CREAMS/WEPP Sediment Deposition Equation: A Semi-theoretical Evaluation', *Trans. ASAE*, **37**(4), 1105–1108.

